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Tsunami Hydrodynamics in the Columbia River

by

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On 11 March 2011, the Tohoku Tsunami overtopped a weir and penetrated 49 km up the Kitakami River, the fourth largest river in Japan (Tanaka et al. 2011). Similarly, the 2010 Chile tsunami propagated at least 15 km up the Maule River (Fritz et al., 2011). In the Pacific Northwest of the United States, large tsunamis have occurred along the Cascadia subduction zone, most recently the 'orphan tsunami' of 1700 (Atwater et al., 2005). The expected future occurrence of a Cascadia tsunami and its penetration into the Lower Columbia River became the subject of "the Workshop on Tsunami Hydrodynamics in a Large River" held in Corvallis, Oregon, 2011. We found that tsunami penetration into the Columbia River is quite different from a typical river. The tsunami enters the vast river estuary through the relatively narrow river mouth of the Columbia, which damps and diffuses its energy. The tsunami transforms into a long period, small amplitude wave that advances to Portland, 173 km from the ocean. Understanding this unique tsunami behavior is important for preparing a forthcoming Cascadia tsunami event.

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The Columbia River

The Lower Columbia River (Fig. 1), with a negligible bed slope of 0.035 m/km, a large tidal prism, and regulated river flows between 2000 m^3/s (late summer) and 17,000 m^3/s (large spring freshet) provides a contrast to steeper, smaller, more confined rivers. The riverbed of the Lower Columbia River remains below mean sea level all the way to Bonneville dam at river kilometer (rkm[*]) 234 (Kukulka and Jay 2003). In contrast, tsunami penetration of the 2011 Tohoku event ceased at the 49 km mark of the Kitikami River, where the bed elevation was 4.6 m above the sea level.

Jay et al. (2011) classified the Lower Columbia River into five reaches. The wave-dominated entrance extends to rkm 5. The lower estuary with extensive salinity intrusion extends to rkm 30. The upper estuary (rkm 30 to 86) is approximately 15 km wide at its lower end, narrows landward, and exhibits complex shallow-water bathymetry with braided channels (see Fig. 1). Moving upstream from rkm 86 to 229, tidal range decreases and water level variations become increasingly dominated by river stage. Finally, there is a steep reach (rkm 229 to 235) where the river flows over the remains of an ancient landslide, upon which Bonneville Dam is built.

Atwater et al. (2005) used tree-ring data to suggest that the Cascadia earthquake/tsunami of 1700 affected Price Island at rkm 56. The 1964 Great Alaskan tsunami was recorded at five tide-gage stations along the river; the largest supra-elevation of 0.25 m at Astoria (rkm 29) decayed to 0.15 m at Beaver (rkm 86), and was detected at rkm 170 in Vancouver, WA (Wilson and Torum, 1972).

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[*] The river kilometer (rkm) represents a location by distance along the river from the ocean.

The 2011 Tohoku Tsunami was detected in Astoria (rkm 29) at 16:00 UTC on March 11, with an initial elevation of approximately 8 cm over tide-level (Fig. 2). After the transition from ebb to flood at 20:00 UTC, the elevation over tide increased to approximately 15 cm. At Skamokawa (rkm 54), no tsunami was detected until the flood tide began at approximately 22:00 UTC. Similar patterns were observed at Wauna (rkm 66) and Longview (rkm 107), and suggest that ebbing flow attenuates the tsunami at the river mouth and estuary. In addition, higher frequency components of the tsunami waves observed at Astoria disappeared in the record at Skamokawa and farther upstream. Evidently, the Tohoku tsunami interacted with the tidal river. Similar spatial patterns and ebb/flood variation were observed for the 1964 Great Alaskan Tsunami (Wilson & Torum, 1964). For reference, the combined river-plus-tidal flow on March 11, 2011 varied between 7000 m^3/s (HW) and 12,500 m^3/s (end of ebb tide) at the USGS gauge (Beaver Camp) at rkm 86.

Simulations of a Hypothetical Tsunami

A hypothetical tsunami was imposed at the river mouth for our simulation exercise performed at the Workshop. The input condition was derived from a synthesized tsunami for the Cascadia event (Priest et al., 2009). Ground subsidence resulting from the fault rupture as well as tsunami-current interactions in the offshore ocean were not considered. The river discharge rate was set constant at 7,000 m^3/sec, and the tidal range at the river mouth is assumed sinusoidal with a period of 12.42 hours, and the amplitude variations of ± 1.5 m. Tsunami simulations were made based on fully nonlinear shallow-water-wave theory and weakly dispersive-wave theory, e.g. Boussinesq-type models.

Our simulation results indicate significant energy dissipation on the over-bank flows in the river estuary. The maximum tsunami height attenuates significantly from 5.6 m at the river mouth to

1.5 m by rkm 29 as shown in Fig. 3. The tsunami attenuation rate decreases in the confined tidal river upstream from rkm 66: the e-folding attenuation distance increased from 22 km in the estuary to 50 km in the river.

Past field observations often found the formation of undular bores[#] in rivers: for example, the 1983 Japan Sea Tsunami (Tsuji et al. 1991), the 2003 Takachi-Oki Tsunami (Yasuda, 2010), the 2010 Chile tsunami (Fritz et al., 2011), and the 11 March 2011 events. In contrast to previous events, our simulations did not show a formation of undular bore in the Columbia River. A likely reason is that when a tsunami enters a large river estuary, short waves riding on the tsunami radiate to the broad shallow-water areas due to diffraction (leaking wave energy normal to the propagation direction). Frictional interaction with the large mean flow preferentially damps short waves. Both processes prevent the formation of undular bores while such bores occur in confined and narrow rivers.

The wave transforms through a transition zone as the river breadth narrows (from rkm 45 to rkm 80), and higher frequency components present at the river mouth are filtered out. The modeled first-pulse at Skamokawa occurs over 3 hours. This behavior of wave period dilation is qualitatively consistent with the tidal gauge records for the Great Alaskan Tsunami and the Tohoku Tsunami (Fig. 2). The waveform transformation is closely related to the bathymetry. The breadth of the river estuary is approximately 10 km at rkm 40, and is comparable with the tsunami wavelength of approximately 15 km. The river estuary is approximately 50 km long, corresponding to two tsunami wavelengths. Tsunami energy propagating along the relatively deep dredged channels cannot be sustained due to diffraction: i.e. lateral energy leakage onto the

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[#] Note that an undular bore is a weak bore (bore = moving hydraulic jump): instead of dissipating energy at the turbulent bore front, the undular bore disperses its energy behind the front.

expanse shallow region. Together with wave scattering by many shoals and islands, tsunami energy in high frequencies is diffused while the energy in low frequencies is dissipated in the lower estuary. Thus, the wave period grows and amplitude decreases by the time the tsunami reaches the upstream river at Beaver (rkm 86).

The foregoing characterization of the river estuary is applicable for tsunamis whose wavelength is comparable to the estuarine spatial scale. For tides that have a much longer wavelength, diffusion does not change the waveform significantly. On the other hand, for swells with much shorter wavelength, wave-current interactions and breaking are dominant at the river mouth, and most of the wave energy does not penetrate into the lower estuary. Tsunami waves represent an intermediate-scale phenomenon that may result in the unique wave transformation in the river estuary.

It is important to recognize that tsunami penetration into the Columbia River is controlled by a combination of convergent bathymetry and bed friction, although high frequency components are controlled by diffusion and scattering in the river estuary. The bed slope is not the dominant factor for controlling tsunami penetration. Vegetation on the riverbanks and islands will be inundated during the tsunami runup in the river estuary. The effects of energy dissipation must be heterogeneous and unsteady because of the variable roughness depending on the water depths and the flow velocities. Furthermore, we found in the workshop that interaction between the tide and the tsunami is important for accurately predicting tsunami penetration effects in the Lower Columbia River.

More detailed outcomes from the workshop can be found at http://isec.nacse.org/workshop/2011_orst/agenda.html

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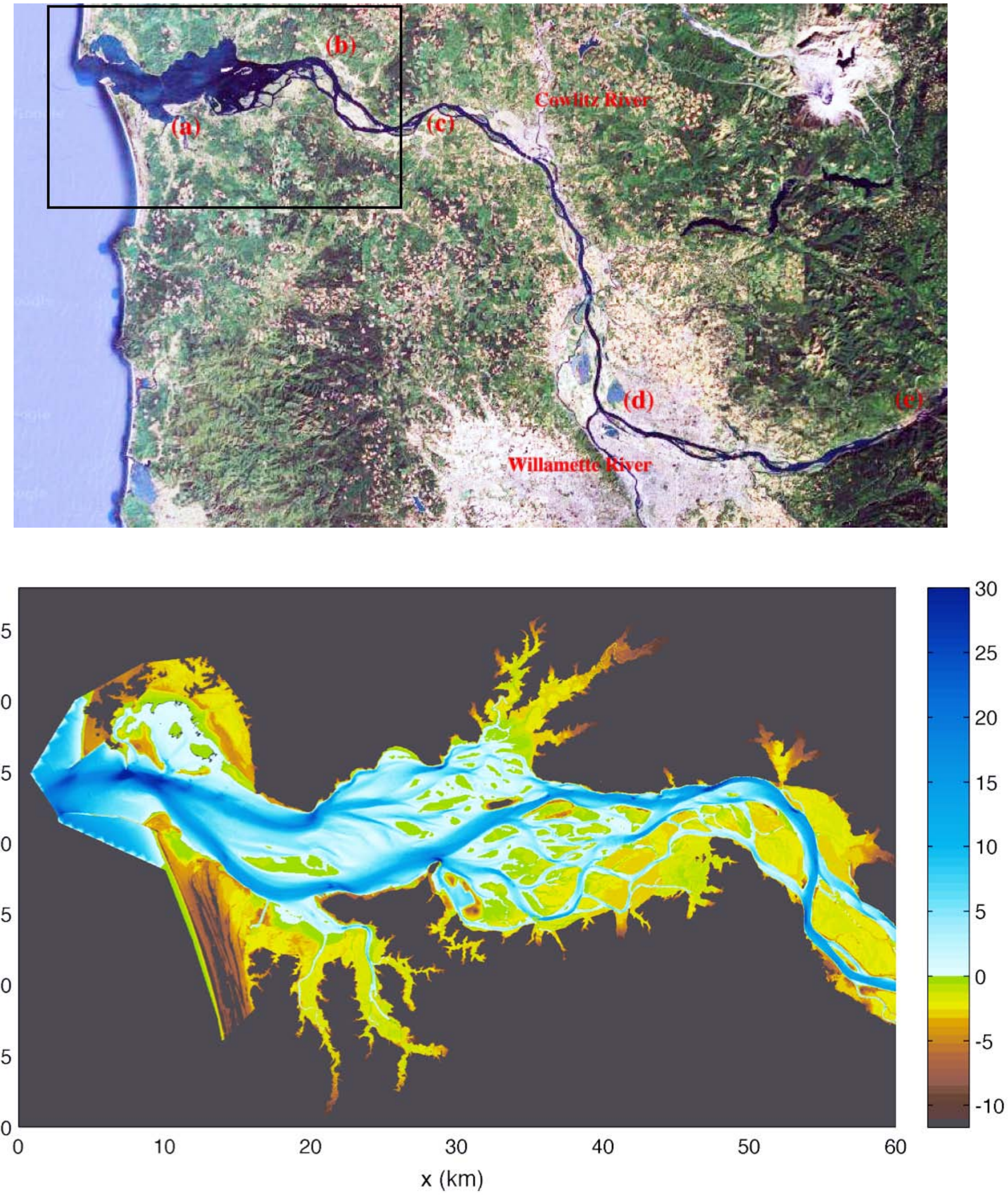

Figure 1. The Lower Columbia River. Top: the river reach up to Bonneville dam (235 km from the Ocean = rkm 235): (a) Astoria (rkm 29); (b) Skamokawa (rkm 54), (c) Beaver Camp (rkm 86), (d) Vancouver/Portland (rkm 170), (e) the Bonneville dam (rkm 235). Bottom: bathymetry of the river estuary (depth in meters) marked in the top map.

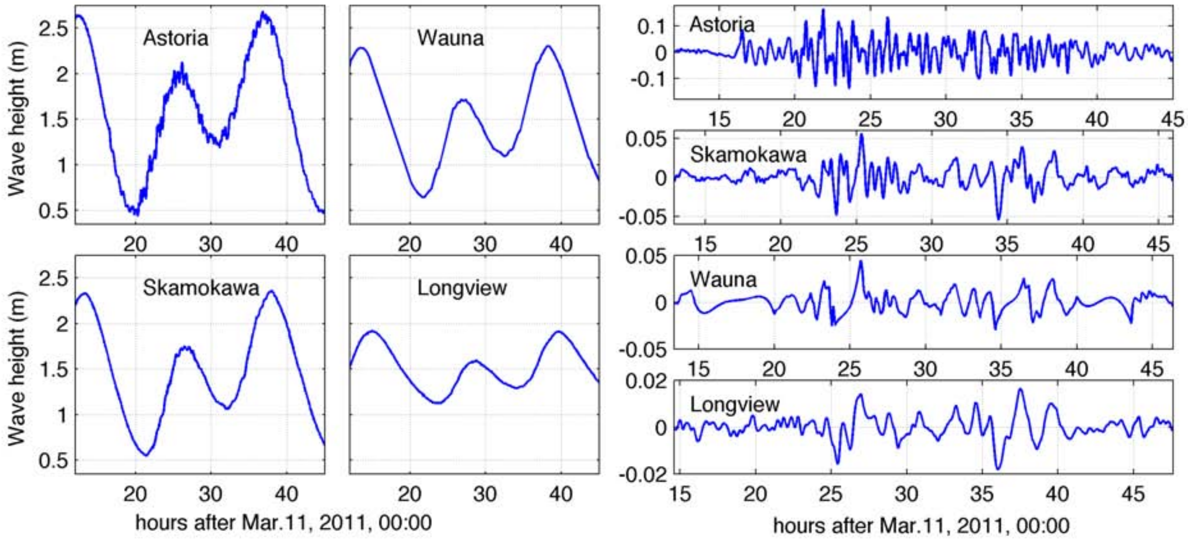

Figure 2. Raw (left panels) and de-tided (right panels) tide gage records of the March 11 2011 Tohoku Tsunami in the Columbia River at Astoria (rkm 29), Skamokawa (rkm 54), Wauna (rkm 66), and Longview (rkm 107). Disturbance propagating with the shallow-water speed would appear on the same vertical line in all the plots of the de-tided records.

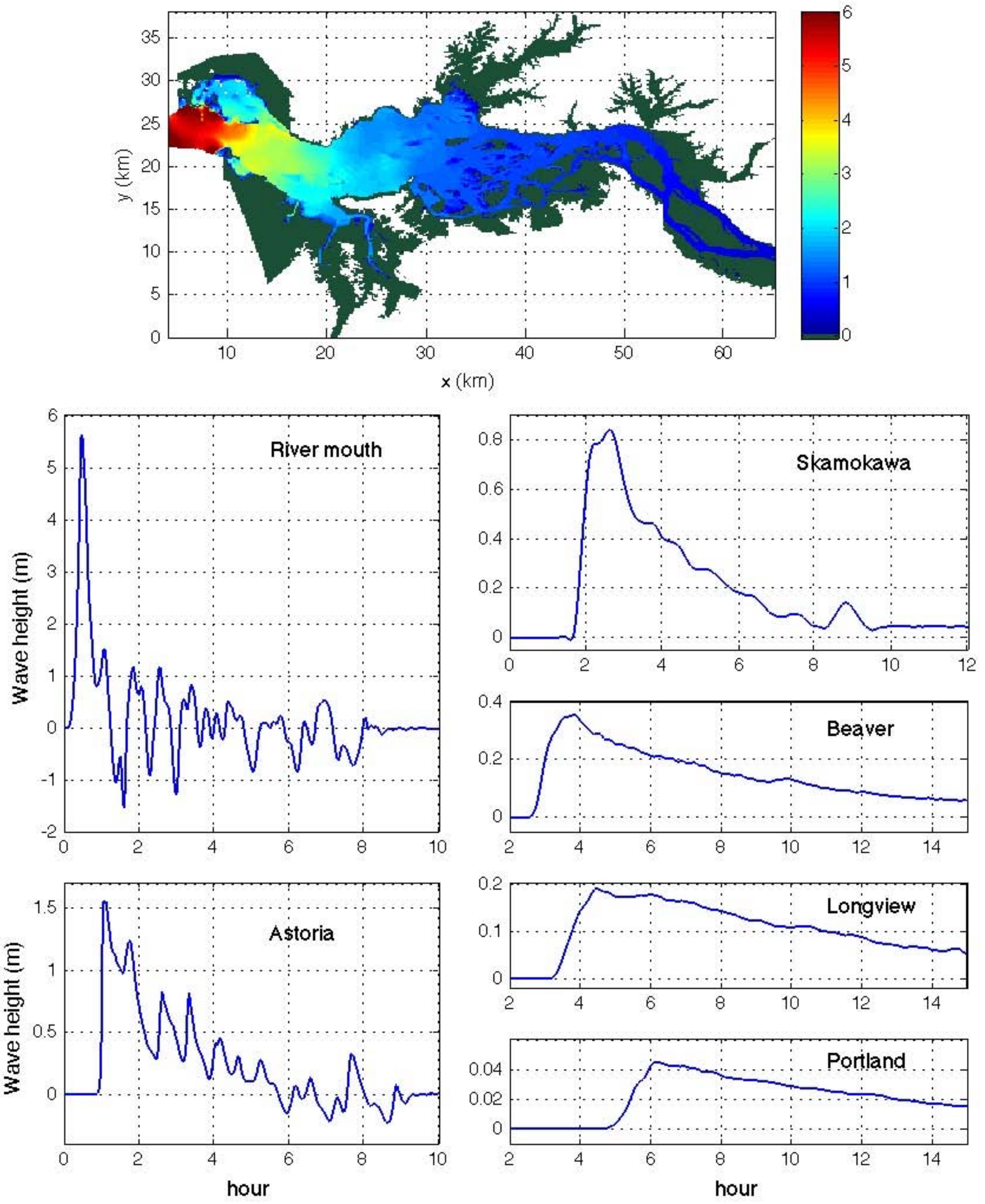

Fig. 3 Numerically simulated tsunami intrusion into the Lower Columbia River. Top: maximum tsunami elevation along the river. Bottom: temporal variations of tsunami forms at five locations: Astoria (rkm 29), Skamokawa (rkm 54), Beaver (rkm 86), Longview (rkm 107), and Portland (rkm 170). Tidal effects are not included for theses simulations for simplicity.